Any Way the Wind Blows

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Chapter 2

The Tundra Ecosystem, Climate and Greenhouse Gas Exchange

2.1 Introduction

In the past century, global temperatures have risen about 0.7 degrees on average, which is commonly attributed to the increase of greenhouse gases in the atmosphere (*Hegerl et al.*, 2007; *Smith et al.*, 2008). Because this is an average temperature increase, it obscures the fact that large differences occur in between regions. For example, certain parts of the Arctic have shown exceptionally high temperature increases of 0.2 to 0.4 °C per decade and this trend is expected to continue (*Zwiers*, 2002; *Chapin III et al.*, 2005; *Serreze and Francis*, 2006). To understand how these temperature changes affect the Arctic as a carbon sink, it is critical to understand the functioning of greenhouse gas exchange in arctic ecosystems. In this chapter, the tundra ecosystem, geomorphology and climate is described, followed by an explanation on the specifics of the exchange of CO_2 and methane in tundra. The chapter concludes with an elaboration on the methods that are typically used to quantify greenhouse gas exchange in the Arctic.

2.2 Vegetation, geomorphology and climate sensitivity

2.2.1 The tundra ecosystem

The tundra climate, according to the Köppen-Geiger classification, falls within the polar climate group under designation ET, which is defined as the area where the temperature of the hottest month is above freezing but below 10 °C on average (*Peel et al.*, 2007). Excluding alpine tundra, this climate region occurs mostly in the most Northern parts of the American and Eurasian continents. Precipitation is quite low in these areas with a typical yearly precipitation of 150 to 250 mm on average, of which most usually falls as rain in summertime and as snow in the rest of the year. Although this amount of precipitation is rather low, total evaporation is also low and thus soils remain wet. Due to the short summers, combined with the low temperatures, tree growth is hindered and therefore most tundra is characterized by low growing vegetation such as shrubs, sedges, grasses, mosses and lichens.
The study area of this research has a highly heterogenous vegetation, which is mostly due to

Figure 2.1: Occurrence of permafrost in the Northern hemisphere. Areas of continuous permafrost are indicated with dark purple. Areas with discontinuous, sporadic and isolated permafrost have been indicated with ever lighter shades of purple. Original image by Philippe Rekacewicz and the International Permafrost Asssociation.

differences in water level. In dry areas, vegetation consists of *Betula nana* and *Salix pulchra* dwarf shrubs or hummocks of the sedge *Eriophorum vaginatum* with *Salix pulchra* in between. In these areas a moss and lichen ground cover is common. In the transition zone from a dry to a wet area, mosses and lichens are replaced by *Sphagnum spp.* and shrub cover is largely reduced. In these parts, *Potentilla palustris* occurs and sedge cover increases to 50% with species such as *Eriophorum angustifolium* and *Carex aquatilis*. These sedges dominate the wet, flooded parts of terrain depressions and polygon centers where *Sphagnum* is absent and sedge cover is around 80 to 100%.

2.2.2 Permafrost and geomorphology

Due to the cold temperatures, most tundra ecosystems are underlain by permafrost. Permafrost is defined as a soil which is below or at the freezing point of water for at least two consecutive years. While permafrost can extend up to several hundreds of meters into the earth, normally the top layer of the soil thaws in summer and this part of the soil, where most soil microbial processes occur, is called the active layer. This active layer can extend up to 2 meters in the taiga, while in tundra ecosystems active layer depths of 30 to 50 cm are more

typical.
The continuous thawing and freezing of the top layer of the soil leads to geomorphological features in the landscape, such as ice wedges, thermokarst lakes and palsas, that are typical for areas with an arctic climate. Ice wedges are created due to thermal contraction cracks in the ground which fill up with water during the summer season. When the water in these cracks freezes, the ice expands and this widens the crack. In following winters, thermal contraction cracks occur in the same places and the process is repeated, thus creating ever expanding ice wedges which also raise the top soil in the process, creating ridges. In arctic regions ice wedges usually join together forming ice wedge polygons in the landscape.
When ice wedges are exposed and melt completely, ponds are created in the surface. The relatively warm water in these ponds leads to a deeper active layer beneath the ponds and if these are deep enough, they do not completely freeze during winter. In the following summer they expand again and the successful succession of this process eventually leads to thermokarst lakes, a common feature of the arctic landscape. The anaerobic conditions at the bottom of thermokarst lakes are perfectly suited for methane production and it has been shown that these lakes are a significant global source of methane (*Walter et al.*, 2006).
A third common feature of the Arctic are palsas, which are small mounds in the landscape. These higher areas are created by frost heaving. This process occurs in areas where water saturated soil freezes, forming an ice lens in the soil. These ice lenses are able to grow larger and larger, provided there is enough water in the surrounding area which they can accumulate. Eventually these ice lenses lift the soil surface, creating frost heaves. These higher areas are more exposed to the winter cold due to a thinner snow cover. This promotes the growth of the ice lens even more, ultimately creating palsas.
While many more periglacial features can be identified, such as pingos and frost boils, the features described above are the most relevant in this thesis. The micro-topographical features created by ice wedges and palsas lead to changes in water availability and thus vegetation. These features therefore govern the spatial variability in greenhouse gas exchange of tundra. Furthermore, ponds and thermokarst lakes are significant emitters of methane, the second most important greenhouse gas.

2.2.3 Sensitivity to climate change

Since global warming occurs two to three times faster in the Arctic compared to the rest of the world, changes to climate change are expected earlier. The currently observed changes are occurring at a rate that has not been witnessed since the end of the last ice age and large effects are observed throughout the Arctic. Sea ice is breaking up earlier and refreezing later, while seasonal minimal extent of sea ice has declined by 45.000 km^2 per year (*Parkinson and Cavalieri*, 2008), furthermore the extent of snow cover has been reduced in the northern hemisphere (*Serreze et al.*, 2000) and these changes are expected to continue.
These changes in temperature and snow cover have a direct effect on the vegetation. Plants flower earlier in the year with earlier snowmelt and changes by as much as 20 days have been reported (*Høye et al.*, 2007). Studies from satellite data have confirmed that the growing season is lengthening across the Arctic and plant growth has increased over the 1980's and 1990's (*Myneni et al.*, 1997; *Stow et al.*, 2004). It has been suggested that this is due to an expansion of shrubs and recently it has been found that this might actually reduce summer permafrost thaw in tundra (*Blok et al.*, 2010). Furthermore, a longer growing season with higher plant activity might lead to more carbon uptake (*Churkina et al.*, 2005).

However, besides these positive effects, higher temperatures also lead to more soil respiration, reducing net carbon uptake (*Euskirchen et al.*, 2006) and seasonal effects might be highly important into explaining differences in the net uptake of carbon (*Randerson et al.*, 1999; *Piao et al.*, 2008). These changes in the net carbon exchange in tundra are important since arctic soils hold large deposits of carbon. Estimates of carbon stores in permafrost regions are currently as high as 1672 Pg or half the amount of carbon stored in soils worldwide (*Tarnocai et al.*, 2009). Recently, it has been shown that carbon released from deeper layers in the soil are also quite sensitive to higher temperatures, increasing the likeliness that respiration can compensate for the higher uptake of carbon due to a longer growing season (*Schuur et al.*, 2008; *Dorrepaal et al.*, 2009). Thus, higher arctic temperatures might lead to less uptake of carbon, which would lead to a higher buildup of CO_2 in the atmosphere and even higher temperatures.

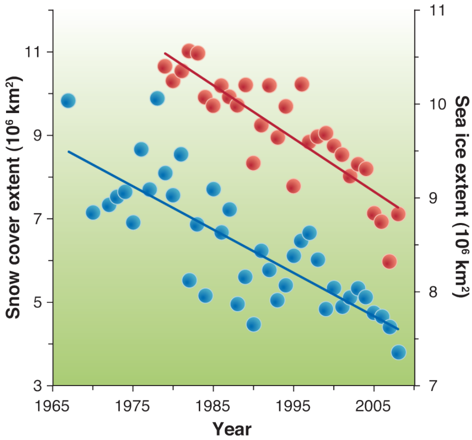

Figure 2.2: Reductions in terrestrial snow cover (blue) and sea ice (red) extent during June to August for the past since the 1960's and 1970's. Original image by Eric Post.

Furthermore, the aforementioned changes in snowmelt and sea ice extent do not only affect arctic flora but can also have large effects on arctic fauna. For example, it has been reported that the migration of caribou to their summer ranges starts due to a change in day length while the flowering of plants is linked to temperature. If the cues to these two processes start to differ in time, these animals can miss the peak of resource availability, which has consequences for successful reproduction (*Post and Forchhammer*, 2008). Furthermore, the reduction in sea ice also puts many marine animals such as seals and polar bears under increasing pressure (*Laidre et al.*, 2008), although recently it has been suggested that greenhouse gas mitigation might be able to help reverse this trend (*Amstrup et al.*, 2010).

2.3 Greenhouse gas exchange in tundra

2.3.1 Photosynthesis and respiration

While human activities release large amounts of CO_2 into the atmosphere, plants take up carbon through photosynthesis. In this process, plants combine water and CO_2 together with sunlight to create sugars, releasing oxygen in the process. The general equation of oxygenic photosynthesis is shown in Equation 2.1

$$2nCO_2 + 2nH_2O + \text{sunlight} \rightarrow 2(CH_2O)_n + 2nO_2 \quad (2.1)$$

The sugars created in this process can be used to create plant tissue or the energy stored within can be used by the plant when and where this energy is needed. This last process is done through cellular respiration and is described in Equation 2.2.

$$C_6H_{12}O_6 + 6O_2 \rightarrow 6CO_2 + 6H_2O \tag{2.2}$$

This process not only releases energy for use by the plant, it also returns CO_2 to the atmosphere. The net carbon uptake of a plant therefore depends on both photosynthesis and plant respiration. Photosynthesis is commonly referred to as gross primary production, GPP, while the difference between GPP and plant respiration is referred to as net primary production, NPP.

Although plants accumulate carbon in their tissue throughout their lifetime, this organic matter is transferred to the soil once plants die or shed leaves. This plant matter is then respired back to the atmosphere by decomposers in the litter layer and the soil, such as bacteria. When considering whole ecosystems it is often difficult to accurately distinguish between this type of respiration and plant respiration. Therefore, usually only the sum of both types of respiration, known as ecosystem respiration, R_{eco}, is discussed. The difference between GPP and R_{eco} is commonly known as net ecosystem exchange, NEE. If GPP is larger than ecosystem respiration, NEE is negative and the ecosystem is taking up carbon from the atmosphere.

Soil temperature and water level are critical in creating the conditions for a net uptake of carbon. Low soil temperatures limit the ability of bacteria to respire carbon and beneath the water table, no oxygen is available for respiration. Areas where large amount of carbon have been stored in the soil are therefore characterized by low temperatures and wet conditions, such as in tundra and peatlands. Due to permafrost melt, these carbon stocks can become more readily available for respiration and in theory these carbon deposits can be respired back into the atmosphere.

2.3.2 Methane

While ecosystems such as tundra and peatlands are usually carbon sinks that have removed large amounts of CO_2 from the atmosphere, the greenhouse gas function of these ecosystems is possibly quite different. Below the water table, where respiration is limited due to the lack of oxygen, organic matter is decomposed by microorganisms known as methanogens, releasing methane in the process. Although this process can follow several pathways, the most commonly known involve CO_2 and hydrogen or acetic acid and are described in Equation 2.3 and Equation 2.4:

$$CO_2 + 4H_2 \rightarrow CH_4 + 2H_2O \tag{2.3}$$

$$CH_3COOH \rightarrow CH_4 + CO_2 \tag{2.4}$$

The methane formed through this methanogenesis has a global warming potential which is 25 times stronger than CO_2 on a per weight basis. Therefore, if enough methane is being emitted, ecosystems such as tundra can be a sink for carbon but at the same time also a source of greenhouse gases in terms of global warming potential. It is therefore critical to understand the processes that determine the emission of methane.

Since methane production only occurs in the anaerobic part of the soil, changes in the water table can have a large effect on emissions, as shown in Figure 2.3. If the top part of the soil is aerobic due to a low water table, methanotrophic bacteria in this part of the soil use oxygen

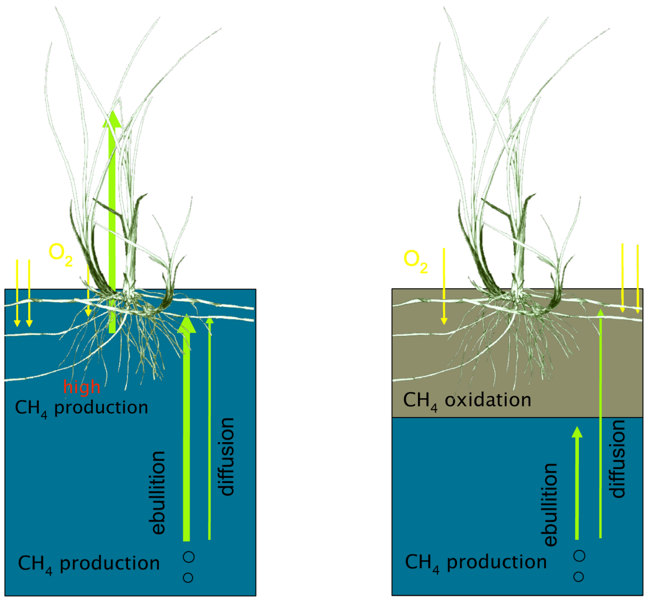

Figure 2.3: Methane pathways in the soil. If water table is high, as on the left, methane (indicated with green arrows) can be transported to the atmosphere directly. If water table is low, as on the right, emissions are limited due to oxidation in the aerobic part of the soil but plants can provide a bypass through this part of the soil. Original image by Ko van Huissteden.

to convert methane into CO_2. Therefore, methane that is produced in deeper soil layers will not reach the atmosphere if the water table is too low.

However, this aerobic part of the soil can be bypassed in certain cases where vascular plants root deep enough to reach the anaerobic part of the soil. Through the aerenchyma of plants, methane can be transported upwards without being oxidized. Therefore, methane emissions are often highest in areas with a large amount of vascular plants (*Joabsson et al.*, 1999; *Greenup et al.*, 2000; *Christensen et al.*, 2003; *Ström et al.*, 2005), although it has been shown that the aerenchyma of plants also provide a pathway for oxygen into the soil, increasing oxidation of methane (*Popp et al.*, 2000; *Whalen*, 2005). Additionally, vascular plants influence the production of methane by providing fresh substrate to the methanogens, which is easier to convert to methane than old and more recalcitrant organic matter (*Ström et al.*, 2005). Since these plants influence methane emissions and because vegetation composition is also related to water availability, vegetation composition can be used as an indicator of the spatial variation of methane emissions (*van Huissteden et al.*, 2005).

Apart from these relationships with vegetation, methane emissions are also increased by higher soil temperatures. In tundra, higher soil temperatures lead to a deeper active layer and this provides a larger reservoir in the soil for methane production to occur. However, higher temperatures also increase evaporation and a deeper active layer possibly increases drainage, which would both influence the water table in the soil. Localized conditions therefore strongly determine the size of methane fluxes and methane emissions are very heterogeneous spatially.

2.4 Measurement techniques

In the past, many different methods have been used to measure fluxes of CO_2 and methane, such as chamber flux measurements, mass balance techniques, eddy covariance, relaxed eddy accumulation and flux gradient methods (*Denmead*, 2008). It is beyond this thesis to discuss all these techniques separately and therefore we limit ourselves to the methods used most in this study, namely the flux chamber method and the eddy covariance technique.

2.4.1 Flux chambers

A widely used method to perform measurements of trace gas fluxes, soil respiration and methane emissions in particular, is the flux chamber technique (*Moore and Roulet*, 1991). This method has the advantage that it's portable and applicable to small areas (< 1 m^2), which makes it ideal to study small scale differences in fluxes. The general application is to place a closed chamber on top of a collar that is placed firmly into the top soil, to avoid leakage to the sides. Subsequently, the concentration increase of the gas of interest within the chamber is measured at regular intervals. By extrapolating the linear relationship between time and concentration, the flux of the gas can be calculated from the volume of the chamber and its base footprint.
The first closed chamber flux measurements were performed as early as the 1960's and 1970's. These early studies used either KOH to absorb CO_2 or they took air samples of the trace gas for analysis in a laboratory afterwards (*Witkamp*, 1966; *Schulze*, 1967; *Kanemasu et al.*, 1974; *Crill et al.*, 1988). Because this practice significantly limited temporal resolution, it was improved upon by performing continuous measurements with an infrared gas analyzer (IRGA) (*Edwards and Sollins*, 1973). This made continuous measurements possible but portability was reduced since a large trailer was needed for the IRGA equipment. After the miniaturization of this technique, chamber flux measurements provided both a high temporal resolution as well as portability (*Vourlitis et al.*, 1993; *Silvola et al.*, 1996).
The flux chamber technique can therefore provide a detailed view on the emission of methane and its spatial variability. If relevant environmental parameters are measured together with the flux measurement, the processes that drive these fluxes can be recognized readily (*van Huissteden et al.*, 2005). However, flux chamber measurements also have many downsides. First of all, measurements are very labor intensive because every plot has to be visited separately. Furthermore, fluxes can be disturbed during the measurement due to, among others, leakage, ebullition (natural or caused by disturbance from the person performing the measurement), heating of the chamber and non-linear concentration increases. Most of these problems can be solved by applying automatic chambers. However, irregular events such as ebullition will still be captured poorly while atmospheric effects such as turbulence are excluded with the closed flux chamber technique. Furthermore, the method is only applicable to low standing vegetation such as grasses and sedges, since chamber sizes are required to be relatively small.

2.4.2 Eddy covariance

Theory

Most of the disadvantages mentioned above do not apply to the eddy covariance technique. This technique can measure trace gas fluxes over an integrated area and continuously. The

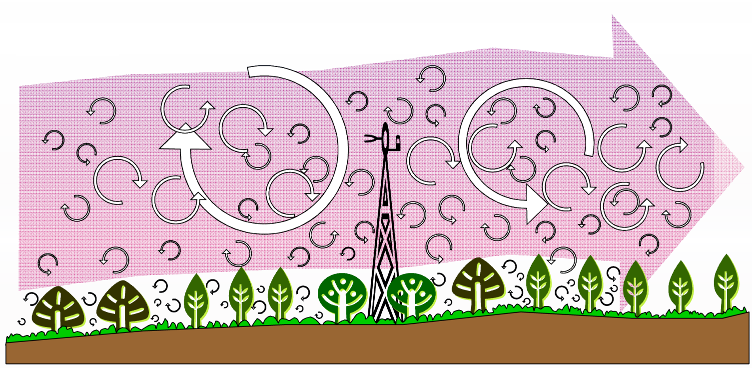

Figure 2.4: On this picture, the air flow is represented by the large pink arrow that passes through the tower and consists of different size of eddies. Conceptually, this is the framework for atmospheric eddy transport. Original image created by George Burba.

method combines measurements of wind speed and concentrations of the trace gas to approximate the average flux through statistical methods. Measurements are performed in open air and therefore atmospheric effects are incorporated. Also, measurements can be performed over larger vegetation such as forests.

Figure 2.4 shows the concept of atmospheric eddy transport. The large pink arrow represents the general airflow while within this airflow many different deviations from the average occur, caused by turbulent eddies. These eddies each have a certain concentration of trace gas and a speed associated and if these concentrations and vertical movements are measured in time, one can determine the flux. For example, if a moving parcel of air has a concentration of s1 and a speed of w1 in an upward direction, while a following parcel of air has a concentration of s2 and w2 in a downward direction, the average will determine how many molecules have traveled up or down and provides us with the flux. In eddy covariance these small eddies are measured as deviations from the average airflow and together with the simultaneously measured deviations of the concentration of the studied gas from the average, the flux can be calculated according to Equation 2.5.

$$F_s = \overline{w's'} = \frac{1}{t_2 - t_1} \int_{t_1}^{t_2} w'(t)s'(t)dt \qquad (2.5)$$

Here F_s is the atmospheric flux of the studied gas (i.e. CO_2, CH_4 or H_2O), w' the deviation of the vertical wind speed from the mean, s' the deviation of the concentration of the studied gas from the mean and t_1 and t_2 the start and end of the studied time period.

Equipment

Since the measured eddies vary at a high frequency, the equipment to measure these fluctuations should measure at a high frequency too. Commonly, equipment measures wind speed and gas concentrations at a minimal frequency of 10 Hz and measurements need to be precise, highly accurate and stable over time. These requirements and data storage issues prevented a general application of eddy covariance before the 1980's (*Lloyd et al.*, 1984; *Businger*, 1986;

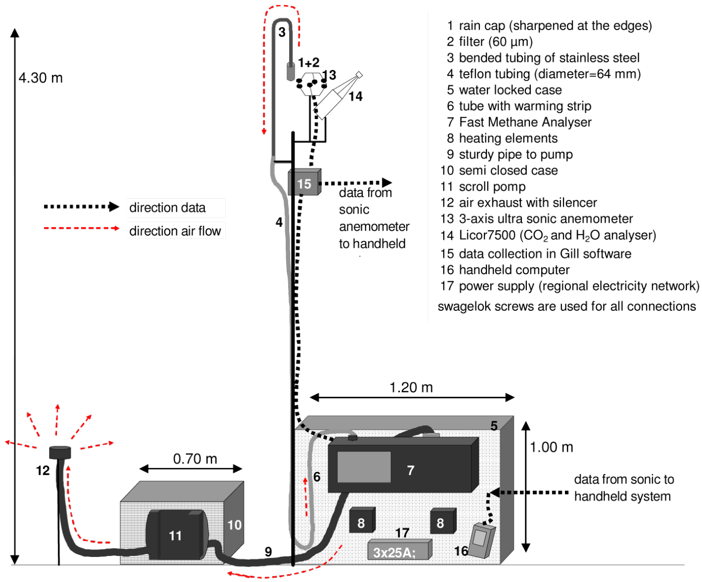

Figure 2.5: Diagram of an eddy covariance tower with a closed path setup for methane measurements and an open path setup for CO_2 and H_2O fluxes. Original image created by Dimmie Hendriks.

Verma et al., 1986; *McMillen*, 1988; *Valentini et al.*, 1991), although the theory was developed as early as the 1950's (*Swinbank*, 1951).

While these early studies used varying techniques, most eddy covariance studies on fluxes of CO_2 or water vapor use a sonic anemometer combined with an infrared gas analyzer (IRGA). A sonic anemometer measures wind speed by measuring the speed of sound between pairs of transducers. Since the speed of sound is also dependent on temperature, it can be derived from these measurements as well. A typical sonic anemometer in eddy covariance studies uses three pairs of transducers which are positioned in such a way that they provide a three dimensional measurement of wind speed and direction.

The IRGA's and other gas analyzers that are used in eddy covariance studies can be divided into two categories: closed and open-path. A closed path setup is often used if the analyzer is too large to place near the sonic anemometer. Instead, an inlet of a tube is situated near the sonic anemometer and from here a gas sample is drawn through this tube towards the gas analyzer. However, the air flow through the tube must be fast enough to replace the air in the measurement cell at the required frequency. Therefore, closed path setups often use a large pump and this increases power requirements.

In an open path setup, the analyzer is much smaller and measurements of gas concentrations are performed in open air. This eliminates the need for a pump and measurements require much less power. However, measurements are more easily disturbed by fog, raindrops on the instrument or radiant heat from the device itself (*Burba et al.*, 2008). This means that additional correction factors are needed and more data has to be filtered out with these types of setups. Even so, the introduction of open-path CO_2/H_2O analyzers such as the Licor-7500 (Licor, Lincoln, NE, USA) greatly simplified eddy covariance setups and made it possible to apply the technique in remote areas with limited power supply.

Since, many studies on CO_2 fluxes have used this equipment over a wide range of terrain but before recently, eddy covariance measurements of CH_4 still had to rely on close-path setups. These setups normally made use of a fast response tunable diode laser spectrometer (TDL) which was shown to give good results (*Verma et al.*, 1992; *Zahniser et al.*, 1995). However, this system had the large drawback that it needed to be cooled, with either liquid nitrogen or a cryogenic system, and that they required regular calibration with a standard gas. These requirements made the system of limited use in remote areas where either liquid nitrogen, calibration gases or the required power are logistically difficult. Recently, an off-axis integrated cavity output spectroscopy method was introduced that does not require cooling or regular calibration (*Hendriks et al.*, 2008). While this method still requires a large pump to obtain the required air flow for high frequency measurements, it is much more applicable in remote areas, provided a stable power supply is present.

Networks and Standardization

Many of the early studies were performed in short campaigns at varying locations with varying techniques. However, this approach makes it difficult to integrate observations to a regional scale or parameterization of models. Therefore, several regional networks (i.e. Euroflux/CarboEurope, Ameriflux and Asiaflux) have been established in the second half of the 1990's to standardize methods, initiate long-term observations and exchange data. Since 1998, these regional networks function under the umbrella organization 'Fluxnet' and currently over 500 long-term tower sites are associated. Through this network large steps towards standardization have been taken. For example, (*Aubinet et al.*, 2000) established the Euroflux methodology that standardized flux calculations of CO_2, H_2O and energy. Also, standard methods for gapfilling and error estimates have been established (*Papale et al.*, 2006). However, these methods mostly focused on CO_2 fluxes and standardization in eddy covariance measurements of methane and post-processing of data is still less defined.

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